Land Surface Temperature Measurements from EOS MODIS Data

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Zhengming Wan, University of California at Santa Barbara
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1. Task Objectives

1) to evaluate the effects of angular variations in the land-surface temperature and emissivity, and thermal infrared BRDF;
2) to investigate the effect of uncertainties in water vapor continuum absorption and to validate the radiative transfer model ATRAD with SST measurements data;
3) methodology development for land-surface temperature and emissivity measurements;
4) to prepare land-surface temperature algorithm and validation plan for peer review.

2. Work Accomplished

2.1. Effects of Angular Variations in Surface Temperature and Emissivity, and Thermal Infrared BRDF

Some modifications have been made in the atmospheric radiative transfer code ATRAD in order to evaluate the effects of angular variations in the land-surface temperature and emissivity, and thermal BRDF (bidirectional reflectivity distribution function). Simulations in the 8-13 μm range show that thermal infrared signals at the top of the atmosphere depend only on the surface temperature and emissivity in the viewing direction under clear-sky conditions if the surface reflects downward sky radiation as a Lambertian surface. This means the contribution from atmospheric scattering of surface thermal radiation in other directions is negligible under clear-sky conditions. Therefore, the single direction measurement of surface temperature by a multiband sensor on satellites can not retrieve the angular dependences of surface temperature and emissivity even if these dependences exist. Cirrus and fogs may bring some cross effects but also make atmospheric corrections more difficult. In real applications of the land-surface temperature estimated from satellite thermal infrared data, surface structure which makes the temperature and emissivity angular dependent may be estimated from visible, near-infrared and medium infrared data. Experience from ground measurements and simulations will be useful to provide some necessary information.

Simulations also show that the effect of thermal infrared BRDF is an important factor to consider in humid regions where the downward sky thermal radiation is comparable to the thermal radiation emitted from the surface. The sea surface is a specular reflecting surface when it is flat under low wind speed conditions. As wind speed increases, the average reflectivity becomes more like a Lambertian surface. If a Lambertian surface has the same spectral and angular emissivity as a calm sea surface which has a specular reflectivity. The difference between the NOAA AVHRR band brightness temperatures for a surface with these two different BRDF patterns could be 0.7-1.0 °K in tropical atmosphere [1].

2.2. Research on Effects of Uncertainties in Water Vapor Continuum Absorption

2.2.1. SST Data Used to Validate Atmospheric Radiative Transfer Models

Thanks to Dr. Ian Barton, a member of the MODIS Science Team, who sent to me a complete data set of the sea-surface temperature (SST) measurements with radiometer and NOAA AVHRR. This data set was used in his paper on infrared continuum water vapor absorption coefficients [2]. The ship data were collected from three different geographical areas around Australia in 1984 and 1985. The data consisted of a clear sky view of the sea surface from an appropriate satellite, a simultaneous ship measurement of SST, and a coincident set of vertical temperature and humidity profiles from a balloon borne radiosonde launched from the ship. The measurement accuracies for temperature and humidity are 0.1 °K and 10%, respectively. The ship-based radiometric SST measurements were corrected for reflected sky radiation using a measurement of sky radiance and an estimate of the surface reflectivity. The SST accuracy was 0.2 °K. The SST data collected in Bass Strait and Coral Sea, as shown in Table 1, are measured with an accurate infrared radiometer so that we are not worried about the difference between sea surface skin temperature and the bulk SST. For a higher confidence in validation of radiative transfer models, we select a most useful sub set, those with data id. numbers d03, d09, d10 and d13 from total 14 sets of data, according to the following criteria: 1) wind speed is less than 10 knots (about 5 meter per second) in order to consider the sea surface as a calm surface. Otherwise, the sea surface BRDF pattern may bring some uncertainties in SST. 2) the maximum humidity is not larger than 95% so that the effects of sub visual cirrus and light marine fog are most likely negligible. Their temperature and humidity profiles are shown in Figures 1 and 2. The humidity profile of data id. d11 is also included in Figure 2. The relative humidity at elevation 8.6km is 98% so it is very possible that there was some cirrus at that time.

TABLE 1. Sea surface temperature measurements data sets, by courtesy of Barton [2].

data id.	date d/m/y	Lat. (°S)	Long. (°E)	zenith angle	SST (°C)	satellite T_4	satellite T_5	wind speed knots/direction	$RH_{surface}$ (%)	RH_{max} (%)	at elev. (km)
Base Strait (NOAA-7)											
d01	05/07/84	40.77	147.92	33°	12.6	9.6	8.7		56	95	1.39
d02	08/08/84	34.88	151.22	59°	18.0	13.7	12.1	18/080	70	76	0.28
d03	08/08/84	36.57	150.35	27°	15.6	13.6	12.6	2/275	58	60	3.39
d04	07/10/84	32.90	153.53	16°	20.7	17.5	16.2		40	93	1.09
d05	12/10/84	31.58	153.90	46°	20.3	18.1	17.1		70	88	0.32
d06	13/10/84	31.75	153.18	49°	21.2	18.0	17.0		72	85	0.01
d07	01/12/84	38.80	148.33	26°	16.1	14.6	13.6	15/090	60	81	1.02
d08	02/12/84	38.67	148.12	3°	15.0	13.5	12.9	25/010	94	94	0.00
Coral Sea (NOAA-9)											
d09	25/10/85	18.42	153.50	50°	26.7	19.9	17.7	8/150	95	95	0.00
d10	28/10/85	15.57	156.53	65°	28.4	15.6	12.6	3/240	70	91	2.73
d11	29/10/85	13.37	154.93	54°	29.2	19.1	16.0	4/160	62	98	8.60
d12	31/10/85	13.03	151.75	36°	27.3	22.9	21.4	13/270	59	105	9.40
d13	31/10/85	13.53	150.75	5°	27.1	22.5	20.7	9/280	59	91	1.86
d14	04/11/85	16.57	147.68	43°	26.9	17.9	15.1	17/175	82	98	5.09

2.2.2. Atmospheric Radiative Transfer Simulations

Because of the high spatial variability in the humidity profile, a great care was taken in selecting elevation levels in radiative transfer simulations, as shown in Figure 3.

The radiative transfer ATRAD [3, 4] has been used for radiative transfer simulations. This model includes following important features: scattering and absorption of different aerosols; different water vapor band absorption models to deal with the temperature and pressure dependences; a sea surface emissivity model to deal with the effect of wind speed and the viewing angle dependence; and different options for water vapor continuum absorption coefficients have also included most recently.

According to the marine aerosol model as used in LOWTRAN7 [5], the density depends on wind speed and a parameter ISCTL (1 for open ocean, 10 for areas under strong continental influence) and the extinction and absorption coefficients of marine aerosols depend on humidity. Because Nass Strait and Coral Sea are close to the Australia coastal line, ISCTL is set to 7. Then marine aerosol densities at selected levels in the boundary range could be calculated using the sea surface wind speed data in Table 1. Background aerosol density was also interpolated at other levels. Above elevation of 13km, the winter midlatitude profile in LOWTRAN7 was used for d03, and the tropical profile was used for d09, d10 and d13.

The sea surface emissivity model [6] was used to calculate spectral emissivities at a series of directions. The specular reflecting pattern is assumed for the sea surface under low wind speed conditions.

The wide used split-window multichannel sea-surface temperature (MCSST) algorithm [7], which was determined by regression analysis of many coincident satellite and ship or drifting buoy measurements, is

$$T_{ss} = 1.0346\,T_4 + 2.5779\,(T_4 - T_5) - 10.05 \tag{1}$$

for NOAA-7 and

$$T_{ss} = 0.9864\,T_4 + 2.6705\,(T_4 - T_5) + 4.24 \tag{2}$$

for NOAA-9. Where T_{ss} is sea surface temperature, T_4 and T_5 are brightness temperatures of AVHRR channels 4 and 5. They are all in degree Kelvin. This MCSST algorithm could be used to evaluate the accuracy of a atmospheric radiative transfer model. In order to validate our radiative transfer model, we also made a more direct comparison between the band brightness temperature from satellite measurements and those from radiative transfer simulations based on measured SST, and atmospheric temperature and humidity profiles.

Detail comparisons are shown in Table 2. The methods to deal with water vapor band and continuum absorptions are indicated in the first and second columns, respectively. The difference between model T_4 and satellite T_4 is given in the third column, and model $T_5 -$satellite T_5 in the fourth column. The difference between model MCSST and SST is given in the last column. The data id., precipitable water, SST and the difference between MCSST from satellite data and SST are given in braces { }. The label ‘‘exptbl (lowtran7)’’ means that the exponential-sum-fitting table for H_2O band absorption is obtained by applying the ‘‘exponential-sum-fitting’’ method [8] to the H_2O transmission values calculated according to the method in LOWTRAN7 [5]. Similarly, ‘‘exptbl (modtran)’’ means the exponential-sum-fitting table from the transmission values according to the method in MODTRAN [9] after a quadratic regression procedure is used to produce the temperature and pressure scaling factors for the effective absorb amount [1].

The label ‘‘lowtran7’’ in the second column means that the H_2O continuum absorption (self-broadening and air-broadening) coefficients are given as in LOWTRAN7, i. e., interpolated or extrapolated from two values at temperatures 296 and 260 °K. Noted that the coefficients themselves at 260 °K were extrapolated from values at higher temperatures from Burch and Alt, 1984 [10]

The label ‘‘exp. form’’ means that an exponential form

$$\ln C_s^0 = \Theta / T + constant \tag{3}$$

is used to express the temperature dependence of the H_2O self-broadening coefficient. This exponential form is supported by the dimer theory and recent measurements by Varanasi [11], and the value of Θ is 2501 °K based on laboratory measurements and 2516 °K based on the dimer theory. In order to keep the consistence between band absorption coefficient and the continuum absorption coefficient as used in LOWTRAN7, we also tried to use the measurement data of Burch. The early measurement data for pure H_2O of Burch and Gryvnak [12] show a very strong negative temperature dependence. The results at three temperatures 296, 392 and 430 °K indicate that the Θ value increases as temperature decreases and as wavenumber decreases. This suggests an approximation

$$\Theta = \frac{5.848 \times 10^8}{\nu T} \tag{4}$$

for the 700-1000 cm^{-1} range, where ν is wave number in cm^{-1}. The H_2O continuum absorption coefficients in LOWTRAN7 are based on new measurement data of Burch and Alt, 1984 [10]. The Θ value calculated from data at 284 and 296 °K is quite close to the value given by Eq. (4). As one option in our radiative transfer simulations, we use the ‘‘exp. form’’ based on the value of C_s^0 at 296 °K in LOWTRAN7 and Θ given by Eq. (4).

The molecular band absorption coefficients used in MODTRAN [9] are calculated at temperatures 200, 225, 250, 275 and 300 °K for any given pressure at a wavenumber interval of 1 cm^{-1}. As shown by Wan and Dozier [13], there are some significant differences between H_2O band models in MODTRAN and LOWTRAN7. In order to use MODTRAN's H_2O band absorption coefficient, the H_2O continuum absorption coefficients may need to modify accordingly. The data of atmospheric transmittance measurements from the Technion Institute in Israel (Oppenheim and Lipson, 1985) were used to validate FASCOD2 and LOWTRAN7 [14]. The measurements were taken with a circular variable filter (CVF) spectrometer for the 8-12 micron window. As shown in Figure 4, the data of Oppenheim and Lipson are compared with simulation results from using H_2O continuum absorption model in LOWTRAN7, H_2O band models ‘‘exptbl (lowtran7)’’ and ‘‘exptbl (modtran)’’. The effects of aerosol and ozone, and uniformly mixed gasses are also included in simulations. It seems that a change of about ±15% for the H_2O self-broadening absorption coefficient may be necessary in the wavelength range 10-12.5 μm. The label ‘‘1.157×exp.form’’ in the second column of Table 2 means that the ‘‘exp. form’’ continuum absorption coefficient is increased by 15.7 percent.

The atmospheric effect on the band brightness temperatures may be expressed by temperature deficit SST$-T_i$, $i = 4, 5$. The values from simulations with different H_2O absorption models are given in Figure 5. It may be easy to make some adjustments for a better agreement in a specific dat set. But a general agreement in all data sets simulated will be more meaningful. Because the viewing angle for d09 and d10 is equal to or larger than 50 °, the simple MCSST may be not suitable. So the numbers in the third and fourth columns and values of model MCSST - SST for d03 and d13 (they are shown in a larger size) are most valuable to validation of radiative transfer models. From Table 2 and Figure 5, we can make following comments on atmospheric H_2O absorption: 1) LOWTRAN7 may overestimate H_2O absorption in dry atmospheric conditions; 2) The band absorption in MODTRAN is less than that in LOWTRAN7; 3) An exponential form of continuum absorption makes model T_4 and T_5 close to satellite data T_4 and T_5 only in cases (d03 and d10). 4) a larger continuum absorption may be necessary to combine with the band absorption in MODTRAN. 5) Because there are significant spatial and temporal variations in the atmospheric water vapor profile, more accurate measurements and simulations are needed before making a defensive conclusion on necessary changes of H_2O absorption coefficients.

TABLE 2. Radiative transfer simulation results compared with SST measurements.

water vapor absorption: band	water vapor absorption: continuum	model T_4-T_4 (°C)	model T_5-T_5 (°C)	model MCSST − SST (°C)
{ data id. = d03, precip. water = 1.33 cm, SST = 15.6°C,				sat. MCSST - SST = 0.45°C }
exptbl (lowtran7)	lowtran7	-0.29	-0.20	-0.10
exptbl (modtran)	lowtran7	0.08	0.30	-0.02
exptbl (modtran)	exp. form	0.03	0.18	0.09
exptbl (modtran)	1.157 × exp. form	-0.05	0.07	0.11
{ data id. = d09, precip. water = 4.13 cm, SST = 26.7°C,				sat. MCSST - SST = -1.20°C }
exptbl (lowtran7)	lowtran7	0.23	0.47	-0.79
exptbl (modtran)	lowtran7	0.60	0.93	-0.66
exptbl (modtran)	exp. form	0.59	0.84	-0.46
exptbl (modtran)	1.157 × exp. form	0.16	0.31	-0.58
{ data id. = d10, precip. water = 4.01 cm, SST = 28.4°C,				sat. MCSST - SST = -4.21°C }
exptbl (lowtran7)	lowtran7	-0.39	0.66	-3.72
exptbl (modtran)	lowtran7	-0.03	1.09	-3.55
exptbl (modtran)	exp. form	-0.25	0.58	-3.02
exptbl (modtran)	1.157 × exp. form	0.02	-0.28	-3.38
{ data id. = d13, precip. water = 3.40 cm, SST = 27.1°C,				sat. MCSST - SST = 0.45°C }
exptbl (lowtran7)	lowtran7	0.22	0.54	-0.21
exptbl (modtran)	lowtran7	0.56	0.96	-0.09
exptbl (modtran)	exp. form	0.56	0.92	0.02
exptbl (modtran)	1.157 × exp. form	0.28	0.55	-0.02

2.2.3. Insights into the Water Vapor Absorption Problem

After a series radiative transfer simulations, we have also gained following insights into the water vapor absorption problem: 1) The viewing angle should be considered in atmospheric correction models; 2) More atmospheric transmittance measurements at a higher spectral resolution should be made to validate radiative transfer models; the resolution 2-6% of the wavelength used in CVF measurements is two low for validation of MODIS thermal infrared bands; 3) The downward sky radiance is more sensitive to changes of absorption coefficients in terms of both values and spectral features, as shown in Figure 6. Therefore, spectral measurements of sky radiance combined with measurements of atmospheric temperature and humidity profiles will be necessary to validate radiative transfer models which are used for development of accurate SST and LST algorithms.

2.3. Methodology Development for Land-surface Temperature and Emissivity

2.3.1. Basic Assumptions in Emissivity Measurements

Most laboratory and field emissivity measurements of terrestrial materials depend on the following basic assumptions: 1) The surface temperature does not change during the TIR measurement or the correlation between the surface temperature and variations in the external radiation source is negligible; 2) The surface emissivity does not change during the TIR measurement; 3) The surface is a Lambertian surface or a specular reflecting surface unless a complete set of bidirectional reflectance is also measured.

Regarding the first assumption, emitted spectral radiance L into direction $\mu = \cos\theta$ at wavelength λ from a surface at thermodynamic temperature T_s is

$$L(\lambda, \mu) = \varepsilon(\lambda, \mu)\,\mathrm{B}(\lambda, T_s) = \varepsilon(\lambda, \mu)\,\frac{2\,h\,c^2}{\lambda^5\,(e^{h\,c/k\,\lambda\,T_s} - 1)} \tag{5}$$

where $\varepsilon(\lambda, \mu)$ is emissivity, $\mathrm{B}(\lambda, T_s)$ is the blackbody radiance at temperature T_s. It is easy to find the relation between the relative change of B and the relative change of the temperature,

$$\frac{d\mathrm{B}}{\mathrm{B}} = \frac{(h\,c/k\,\lambda\,T_s)\,e^{h\,c/k\,\lambda\,T_s}}{(e^{h\,c/k\,\lambda\,T_s} - 1)}\,\frac{dT_s}{T_s} = \eta(\lambda, T_s)\,\frac{dT_s}{T_s} \tag{6}$$

The quantity η decreases with the temperature increase and the wavelength increase. For example, η(3.75 µm, 300 °K) = 13.2, η(3.75µm,1000°K) = 3.92, η(11µm,300°K) = 5.23, η(11µm,1000°K) = 1.79. The relative change of B is 4.34% at 3.75 µm and 1.73% at 11 µm as T_s changes by 1 °K at 300 °K, or 0.39% at 3.75 µm and 0.18% at 11 µm as T_s changes by 1 °K at 1000 °K. This means that the same amount of temperature change, 1 °K, will have a much larger effect on deriving surface emissivity from TIR measurements at T_s 300 °K than at T_s 1000 °K. Usually, it is difficult to check the surface temperature change during real measurements. So we have to pay great attention to the first assumption in measurements made at a relative low surface temperature.

Regarding the second assumption, although the surface emissivity usually does not change during measurements for inorganic samples, it may be a problem for measurements of terrestrial organic samples such as soils, vegetation and tree leaves, because their moisture conditions may change during measurements, especially if measurements are made at high temperature conditions or if a powerful external radiance source is used in order to obtain a high signal-to-noise ratio. However, we can easily check this assumption by repeated measurements. In order to avoid emissivity variations, it is better to use a low-power external radiance source and to make many infrared measurements under natural conditions in the field. Then the averaged TIR data should be used to derive the surface spectral emissivity.

Regarding the third assumption, the best way is to make bidirectional reflectance measurements for a given land cover with an infrared goniometer at the wavelengths of interest in order to check whether the surface can be approximated by a Lambertian surface. In most cases, we use this assumption first, and then check its suitability by changing the incident direction of the external radiation source.

2.3.2. A Four-Step-Method for Field Measurements

Some thermal infrared spectral radiometric measurements has been made at a grass land by the Lewis Research Center baseball field under clear-sky conditions in early Spring of 1992. Measurements were also made over a concrete surface. We used the solar beam as an external radiation source in the following four-step method: step 1, measure the sample surface under sunshine; step 2, measure the sample surface under a shadow by blocking the solar beam; step 3, measure the diffuse gold plate surface under sunshine; step 4, measure the diffuse gold plate surface under a shadow by blocking the solar beam. Use of the solar beam as an external radiation source has the following advantages: 1) It is natural and makes measurements easy; 2) The solar beam is a major part (> 96%) of the total solar radiative flux to the Earth surface in the wavelength range 2.5-14.5 µm under clear-sky conditions; 3) The surface emissivity of the land cover usually does not change when it is moved from sunshine to shadow or the reverse; 4) The solar beam at the surface is only effective for wavelengths shorter than 4.2 µm and it is negligible for 8-14.5 µm. Therefore, the environmental radiation (including solar and sky radiations) in 8-

$14\,\mu m$ exposed on the sample surface does not change after the solar beam is blocked. So the radiance from the surface only changes with the surface temperature. As a result, the change of surface temperature may be well estimated from the radiance change.

The measured radiance from land surfaces changes with time due to the random noise of the measurement system, surface temperature change, and stochastical changes in the atmospheric state. In the atmospheric windows 3.4-4.1 and 8-13μm, the first two reasons are the dominant factors. A temporal analysis of the concrete surface TIR data at wavelengths 3.75 and $12\,\mu m$ shows radiance standard deviations, $dL(\lambda)/L(\lambda)$, of 2.3% and 0.2% respectively under sunshine, or 5.8% and 0.3% under shadowing. According to Eq. (6), $\eta(3.75\mu m)/\eta(12\mu m) \approx 2.5$, but the ratio between the measured dL/L at $3.75\,\mu m$ and the measured dL/L at $12\,\mu m$ is larger than 19 in the shadowing condition. It is estimated that the TIR instrument has a lower signal-to-noise ratio, about 20, at the shorter wavelength, $3.75\,\mu m$. A higher signal-to-noise ratio could be achieved by averaging many spectra, or equivalently, by using a slower scan speed. During TIR spectral measurements of the grass field, we used a broad band (8-14$\,\mu m$) infrared thermometer to monitor the temporal change of the surface temperature at a display rate of 2 per second. We found that the grass field surface temperature may change quickly by 0.2-0.5 °K due to variations of the surface wind speed. The average wind speed was estimated at 1-2 m/s during our measurements. This also suggests the necessity of averaging measurement data. At least 16 spectra were averaged before deriving surface temperature and emissivity values from TIR measurement data.

Now we consider how to estimate a change of the average surface temperature that occurs when a land surface is moved from sunshine to shadowing, or the reverse. We can estimated the surface temperature change by using its brightness temperature as follows. According to the definition, brightness temperature T_b is related to the surface emissivity and temperature by

$$\mathrm{B}(\lambda, T_b) = \frac{2hc^2}{\lambda^5 (e^{hc/k\lambda T_b} - 1)} = \varepsilon(\lambda)\,\mathrm{B}(\lambda, T_s) = \varepsilon(\lambda)\,\frac{2hc^2}{\lambda^5 (e^{hc/k\lambda T_s} - 1)} \tag{7}$$

It is easy to find the relation between the change in T_b and the change in T_s,

$$\frac{dT_b}{T_b} = \frac{T_b}{T_s}\,\frac{e^{hc/k\lambda T_s}}{e^{hc/k\lambda T_s} - 1 + \varepsilon(\lambda)}\,\frac{dT_s}{T_s} \tag{8}$$

And similarly, the relation between the change in T_b and the emissivity change is

$$\frac{dT_b}{T_b} = \frac{k\lambda T_b}{hc}\,\frac{e^{hc/k\lambda T_s}}{e^{hc/k\lambda T_s} - 1 + \varepsilon(\lambda)}\,\frac{d\varepsilon(\lambda)}{\varepsilon(\lambda)} \tag{9}$$

Because the atmospheric transmission function at $11\,\mu m$ for a path of several meters is very close to 1 and the emissivity of most land covers at this wavelength is larger than 0.95, the error in estimations of the surface temperature change by using the brightness temperature is less than 3%. Thus, the estimated surface temperature change is 3.3 °K for the grass field, or 0.5 °K for the concrete surface. Then this estimated surface temperature change can be used to correct the temperature change effect on surface emittance in the range 3.4-4.1$\,\mu m$. After this correction, the change in TIR data is solely due to passing or blocking the solar beam. We use LOWTRAN7 to calculate solar beam spectra based on date and time, upper air temperature and humidity profiles, and surface air temperature and humidity measurement data. The simulated spectra are expected to have an accuracy of about 5% in this wavelength range [13]. Then surface albedo, $R(\lambda)$, can be estimated from averaged TIR signals obtained under sunshine and shadow conditions under the Lambertian approximation. The relative accuracy of estimated surface albedo is expected to be around 5%. Then the surface emissivity can be calculated by Kirchhoff's law $\varepsilon(\lambda) = 1 - R(\lambda)$. The accuracy of the estimated emissivity is about 0.25-0.5% for the grass field, or 2-3% for the concrete surface if they are good Lambertian surfaces. We expect that the error associated with the Lambertian assumption will be equal to or larger than the others, so a complete set of bidirectional reflectance measurements is needed for accurate estimation of surface emissivity in the field. After the surface reflectivity and emissivity in 3.4-4.1$\,\mu m$ are estimated by using the above method, an external radiation source effective in 2.5-14.5$\,\mu m$ can be used in field measurements. Its incident radiative flux onto the target surface can be determined by directly measurements with a TIR spectral radiometer and a reflecting mirror, or a calibrated standard reference surface (for example, a diffuse reflecting gold plate). The TIR measurement data in 3.4-4.1$\,\mu m$ from the target surface can be used to estimate the surface temperature change ΔT_s, then the estimated

ΔT_s can be used for deriving surface albedo (reflectivity, in general sense) and emissivity at other wavelengths with the same method.

Use of the solar beam as an external radiation source and the four-step method in field TIR measurements has given some encouraging but preliminary results [3]. This procedure may be used to investigate the effects of surface structure and shadowing.

2.4. Land-surface Temperature Algorithm and Validation Plan

We have identified following key issues in the development of LST algorithms [1].

2.4.1. Validation of Atmospheric Radiative Transfer Models

The development of LST algorithms will heavily depend on atmospheric radiative transfer simulations. Validation of the accuracy of radiative transfer models will be a key issue at the highest priority. To make accurate spectral measurements of the atmospheric downward radiation [15, 16] at a spectral resolution 1-5 cm^{-1} in the wavelength range from 3 to 14 μm at several zenith angles may be a practical and yet efficient way for ground validation. Upper air temperature and humidity profiles from balloon borne radiosonde and surface air temperature and humidity data will be used as inputs to radiative transfer models. The ground measurement data of downward spectral sky radiances will be used to validate the results from radiative transfer simulations. These data will be also used together with surface emittance data to obtain spectral reflectivity and emissivity of land covers.

2.4.2. Database of Surface Reflectivity and Emissivity Characteristics

Spectral emissivities of land covers are needed not only for accurately estimation of LST from space, but also for calculation of the total long wave radiation from land surfaces for surface energy budget studies after temperature has been estimated. Salisbury et al. [17] recently published a book on infrared (2.1-25 μm) spectra of minerals, with digital data on CD-ROM. A few spectral emissivity figures of soil and sand samples can be found in open literatures [18, 19]. But development of accurate LST algorithms will depend on improvement of our knowledge on spectral emissivity of natural land covers in the next a few years. We will look into the land cover grouping problem in terms of reflectivity and emissivity characteristics. If some useful relations between visible/near-infrared spectral features and thermal infrared features for various land covers could be found, visible and near-infrared MODIS bands will be used to find land cover groups, each having similar emissivity spectral features. Then relative stable spectral emissivity characteristics in two or more MODIS thermal infrared bands for a given land cover group will be used to develop a LST algorithm specific for that group. Because surface emissivity may change with surface conditions such as moisture [20], it is important to include effects of surface ecological status and environmental conditions on emissivity variations into the database.

2.4.3. A Practical Multiband LST Algorithm Hierarchy

Theoretically speaking, we can develop physical determinative inverse models to estimate LST through forward radiative transfer simulations if atmospheric temperature and water-vapor profiles are accurately known by other means. But actually, accurate determinations of these atmospheric profiles, especially in the surface boundary region which has a significant effect on LST algorithms, depend on surface conditions including temperature, and emissivities in a wavelength range. Therefore, statistical regression models of LST algorithms based on results from systematic radiative transfer simulations over wide ranges of atmospheric and surface conditions may be more practical [3, 4]. In order to achieve the specified LST accuracy, a multiband LST algorithm hierarchy will be developed for different land cover groups with similar emissivity spectral features and for different regions and seasons. The viewing angle will be included in these LST algorithms. Reasonable overlaps should be provided for these algorithms between different regions and seasons. This means that we have to consider a wide range of atmospheric changes in radiative transfer simulations for each algorithm. It is expected that LST algorithms have a similar form but use different regional and seasonal dependent coefficients for different land cover groups. Better knowledge of atmospheric conditions, either empirical information or estimated information from other MODIS bands helps to select a more suitable algorithm from this hierarchy to achieve a high accuracy. If the surface at a given pixel is well known such as in application of SST algorithms, nighttime models are better because of avoiding the difficulty in solar radiation correction. But nighttime thermal infrared data alone are usually not enough to identify or to discriminate land covers. If daytime visible and near-infrared data are combined in use of nighttime thermal infrared data to estimate land cover type and to estimate LST, errors in data co-registration will bring uncertainties into land cover classification and LST estimation. This may causes a significant LST error near boundaries between different land cover groups, but it is not a problem for pixels within large areas of land covers. Daytime LST models using simultaneous visible, near-infrared, and thermal infrared MODIS data avoid the co-registration problem, but face the serious problem of solar radiation correction if the medium wavelength thermal bands in the range 3.4-4.1 μm are also used in LST algorithms. As a first step, the three bands in the range 8-13 μm could be used to estimate LST for land covers which are classified by using spectral information from other MODIS bands. In the next step, three bands in the range 3.4-4.1 μm could be used in surface structure and shadow analysis, and even in estimation of LST. Better LST algorithms will be first developed for some reference land surfaces that uniformly cover a relative large area (at least 10 km by 10 km) where spectral emissivity and reflectivity characteristics are well known and topographic features are relatively simple. Some inland lakes, snow and ice covers in flat areas, uniform forest areas and agriculture field are potential candidates for the reference land surfaces. After land cover classification and temperature estimation for each pixel using daytime MODIS data, the same spectral characteristics of surface emissivity at the pixel, and empirical knowledge on diurnal variation of the surface emissivity for the belonging land cover group, if any, will be used in the nighttime LST algorithm to determine night LST.

2.4.4. Validation and Quality Control of LST Algorithms

High quality ground-based and aircraft measurements will be conducted in various sites of the world to validate the LST product in a way similar to evaluating atmospheric correction models for retrieving surface temperatures from AVHRR over a tallgrass prairie [21]. The quality and accuracy of the LST product depend on successful discrimination between clear sky and cloud covers, and detection of optically thin cirrus clouds [22] The estimated LST accuracy also depends on the confidence level of our knowledge on surface covers: type, reflectivity and emissivity characteristics, and their temporal changes, heterogeneity, topographic features and meteorological conditions. The more we know the surface, the more accurate LST we can estimated from satellite data and the more useful the LST will be. If we know nothing about surface properties and conditions, LST alone is not very useful even if it is accurate. Reasonable regions of multiband temperature deficits will be established by systematic radiative transfer simulations for quality assessment and control.

2.4.5. Close Synergism with Other EOS Data and Products

The LST algorithm will use MODIS thermal infrared band data directly, and other EOS instrument data indirectly. ASTER thermal infrared data with 90-m resolution will be used for spatial analysis and quality assessment of LST. As one of the four core MODIS land products, LST has a very close relation with other MODIS land products, and it will use information from land cover [23], vegetation index [24], snow cover [25] and BRDF studies [26]. It also uses, directly or indirectly, MODIS atmospheric products including cloud, aerosol, water vapor properties [22].

3. Anticipated Future Actions

1) to continue on validation of the radiative transfer model ATRAD in the thermal infrared range from 3.4 to 13μm;
2) to purchase a TIR spectroradiometer in early 1993 so that field measurements of land-surface emissivity and sky radiance could be conducted as soon as possible;
3) to prepare LST algorithm and validation plan for a series of peer review.

4. Publications

Z. Wan, D. Ng and J. Dozier, Spectral emissivity measurements of land-surface materials and related radiative transfer simulations, accepted by *Advances in Space Research*, 1992.

J. Dozier and Z. Wan, Development of practical multiband algorithms for estimating land-surface temperature from EOS/MODIS data, accepted by *Advances in Space Research*, 1992.

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